environment

The science of climate change

Global and U.S. Perspectives

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NATIONAL CENTER FOR ATMOSPHERIC RESEARCH

PEW CENTER on Global CLIMATE CHANGE

Contents

Foreword *Eileen Claussen, Executive Director, Pew Center on Global Climate Change*

This report on the science of climate change seeks to explain how climate is influenced by anthropogenic factors. Understanding the effect of greenhouse gas concentrations on the atmosphere is key to understanding the potential magnitude of the "greenhouse effect," evaluating possible environmental impacts, and considering policy responses.

A variety of factors determine the rate and magnitude of climate change, including the emissions of greenhouse and aerosol-producing gases, the carbon cycle, the oceans, biosphere, and clouds. As our understanding in each of these areas evolves, it is important that researchers, policy-makers, the press, and the public be kept informed since these developments affect our understanding of the seriousness and complexity of this issue.

As part of the Pew Center's series examining the potential impacts of higher atmospheric concentrations of greenhouse gases on the United States, this paper by the distinguished climate scientist Tom M.L. Wigley, senior scientist with the National Center for Atmospheric Research, addresses what is known and not known about the science of climate change. Its publication comes in an interim period between assessments of the science by the Intergovernmental Panel on Climate Change (which published its second assessment in 1996 and will publish its third assessment in 2001). The author uses preliminary estimates of greenhouse gas and sulfur dioxide emissions from the current IPCC review process as well as his own work to supplement previously published research.

The new research suggests the likelihood of slightly larger changes in temperature and sea level rise than projected in the most recent IPCC assessment. The temperature rise is expected to be greater in the U.S. than the average temperature increase across the globe. While changes in precipitation and extreme weather events such as hurricanes and other storms are more difficult to predict, it is possible that the intensity of rain and hurricane events could increase. Uncertainties in predicting the direction and magnitude of these changes make it difficult to predict the impacts of climate change. However, even small changes in climate can lead to effects that are far from trivial.

While the analysis presented is the work of one author, this report has been subject to extensive peer review. The Pew Center and the author are indebted to many scientists and organizations for their constructive comments on previous drafts of this paper or sections of this paper. Their comments have helped improve the text substantially, and so, while the opinions expressed in this report are solely those of the author, we gratefully acknowledge their input: E. Barron, B. Felzer, C. Hakkarinen, A. Henderson-Sellers, M. Hulme, M. MacCracken, M. McFarland, J. Mahlman, G. Meehl, N. Nakićenović, B.D. Santer, M.E. Schlesinger, K.P. Shine, J.B. Smith, and S.J. Smith. The A1, A2, B1, and B2 scenarios developed in the current IPCC working group process have been used with the kind permission of their producers, represented by T. Morita, A. Sankovski, B. deVries, and N. Nakićenović. D. Viner of the Climate Impacts LINK Project (UK Dept. of the Environment, Regions and Transport contract EPG1/1/68) supplied the HadCM2 data on behalf of the Hadley Centre and UK Meteorological Office. In addition, the Pew Center would like to acknowledge and thank Joel Smith and Brian Hurd of Stratus Consulting for their management of this Environmental Impacts series.

Executive Summary

The average surface temperature of the globe has warmed appreciably since the late 1800s, by about 0.6°C. Since this warming cannot be adequately explained by natural phenomena such as increased solar activity, human-induced increases in greenhouse-gas concentrations appear to be at least partly responsible. In addition to the warming effect of greenhouse-gas increases, however, changes in temperature over the past century are likely to have been significantly influenced by the cooling effect associated with changes in the sulfate aerosol loading of the atmosphere, arising from fossil-fuel-derived sulfur dioxide (SO_2) emissions. When greenhouse-gas, sulfate aerosol, and solar influences are considered together, observed climate changes are consistent with model predictions.

Projections of future global-mean temperature and sea level change made by the Intergovernmental Panel on Climate Change (IPCC) in its 1996 Second Assessment Report used emissions scenarios developed in 1992. Preliminary versions of new emissions scenarios produced by the writing team for the IPCC Special Report on Emissions Scenarios (SRES) are now available. The most important difference between the old (1992) and new (SRES) scenarios is that the new scenarios have much lower emissions of sulfur dioxide. The reduction in sulfur dioxide emissions (and their attendant cooling effects through the production of sulfate aerosols) results in a slight increase in temperature and sea level rise projections from those previously given by the IPCC. If central estimates of model parameters are used, global-mean warming from 1990 to 2100 ranges from 1.9°C to 2.9°C. Sea-level rise estimates over the same period range from 46 to 58 cm. For temperature and sea level changes over the next few decades, projections are virtually independent of the emissions scenario.

Based on results from a number of climate models, the rate of future warming over the United States is expected to be noticeably faster than the global-mean rate. Future regional-scale precipitation changes are highly uncertain. The only result that is common to all climate models is an increase in winter precipitation in northern latitudes, from the northern Great Plains to the northeastern states. Even in the absence of large precipitation changes, there could still be significant changes in the availability of water for agriculture, human consumption, and industry because of the increased evaporation that should accompany warming. This factor alone would lead to drier summer soil conditions and reduced runoff. The effects of increased evaporation, however, may be partly offset by the direct plant-physiological effect that carbon dioxide (CO_2) has in improving plant water-use efficiency and, hence, lowering evapotranspiration rates.

Changes in weather and climate extremes over the United States are certain to occur as the global climate changes. The frequency of extremely hot days is almost certain to increase, and the frequency of frosts should decrease. Changes in the frequency of daily precipitation extremes are highly uncertain, although there is evidence for an increase in the frequency of wet extremes. For hurricanes and tropical storms, the evidence suggests that there could be small increases in their windspeeds. It is also likely that future such storms will be accompanied by larger rainfall amounts. While there is no credible model-based information on changes in the number of hurricanes and tropical storms per year worldwide, there is empirical evidence that suggests that a small increase in frequency is possible in the North Atlantic region. For all extreme events, however, it is unlikely that the projected changes will become evident in a statistically convincing way for many decades, with the exception of temperature extremes, which should become evident sooner.

I. Introduction

Both in common usage, and as a technical term, the word "climate" refers to the average weather of a place or region — hot or cold, wet or dry, calm or stormy, etc. In other words, the climate is what one obtains if weather conditions during a particular time of the year are averaged over a number of years. Until recently, changes in climate occurred solely through natural processes, associated, for example, with changes in the output of the sun or slow changes in ocean circulation. Today it is clear that human influences, particularly those related to energy use and land-use changes (including deforestation), can also change the climate.

The human activities that lead to climate change have grown substantially over the past century. Because they are expected to grow even more rapidly in the future, their climate influence is also expected to grow. Indeed, model projections show future changes that are unprecedented relative to past human experience, in terms of both the magnitude and the rapidity of change. The consequences of such changes in climate for humans and the natural environment are potentially serious. It is therefore important to obtain the best possible information about the magnitudes and rates of future climate change (both in the absence of policies to reduce such changes and in response to such policies), and about the impacts of future climate change on human and environmental systems.

In 1988, the Intergovernmental Panel on Climate Change (IPCC) was established to provide reliable, policy-relevant, and policy-neutral information on these and other aspects of the climate change issue.[1] Since then, the IPCC has produced a number of major reports on the subject. The purpose of the present report is to summarize the state of knowledge about climate change both at the global scale and over the continental United States. It draws heavily on IPCC reports, updating this information where necessary.

The primary (but not the only) cause of human-induced or "anthropogenic" climate change is the enhanced greenhouse effect (see the Appendix for further technical details). The word "enhanced" is important here because there is also a natural greenhouse effect. The natural greenhouse effect is the warming of the Earth's surface and lower atmosphere that arises because of the presence of certain gases in the atmosphere — in the absence of these gases the Earth would be much colder than it is at present. The particular gases that lead to this warming, the so-called "greenhouse gases," include carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), ozone (O_3), and water vapor. By adding CO_2, CH_4, and N_2O to the atmosphere — or by adding other greenhouse gases that do not occur naturally, like the halocarbons — we enhance the natural greenhouse effect and cause additional warming. This warming, in turn, increases evaporation rates and adds more water vapor to the atmosphere, causing further warming.

There are two leading climate change questions: how much additional warming will occur if we increase the atmospheric levels of greenhouse gases, and what will the consequences be for the climate system as a whole. These are particularly pressing questions because both the climate system and the carbon cycle, which dominates the human influence on climate, have such large inertia: they are like giant flywheels, slow to start moving, but difficult to stop. What we do to the atmosphere today will continue to affect the climate decades or even centuries into the future, and the efforts we make now to reduce the magnitude of future change will only become apparent slowly, also on timescales of decades to centuries.

In this report, Section II reviews past changes in atmospheric composition and climate. Section III deals with the role that human influences have played in climate change to date. This is the "detection and attribution" issue: have we detected (in a statistical sense) any unusual changes in climate, and, if so, can we attribute these to human activities? Section IV gives projections of future global-mean changes in temperature and sea level. This section describes the preliminary IPCC SRES emissions scenarios and the projected future concentration changes for CO_2. The implications for global-mean radiative forcing, temperature, and sea level are then considered. Section V interprets these global-mean results in terms of their consequences for regional-scale temperature and precipitation changes over the United States, after first assessing the credibility of the models used to obtain these projected changes. Section VI provides an assessment of potential changes in other climate variables and in the frequency of extreme events.

II. Observed Changes

This section reviews observed changes in climate and in the factors that may be responsible for these changes. The main concern is the human influence on climate. Other factors are also considered, since these form the backdrop against which human influences are imposed. The focus is on the largest spatial scales, from the continental to the global, since it is only on these scales that we can currently hope to observe human influences.

A. Changes in Atmospheric Composition

The composition of the atmosphere has changed markedly since pre-industrial times: CO_2 concentration has risen from about 270–280 parts per million by volume (ppm) to over 360 ppm today, CH_4 has risen from about 700 parts per billion by volume (ppb) to over 1700 ppb, and N_2O has increased from about 270 ppb to over 310 ppb. Halocarbons that do not exist naturally are now present in substantial amounts. The pre-industrial levels of these gases are known because the composition of ancient air trapped in bubbles in ice cores from Antarctica can be measured directly (Etheridge et al., 1998; Güllük et al., 1998). These ice cores show that the changes since pre-industrial times far exceed any changes that occurred in the preceding 10,000 years.

Human activities — fossil-fuel burning, land-use changes, agricultural activity, the production and use of halocarbons, etc. — are the dominant cause of these changes. This is undeniable for halocarbons like CFC11 and CFC12 because these gases do not occur naturally. For CO_2, CH_4, and N_2O, the human role is virtually certain too, partly because of the rapidity of changes since pre-industrial times, but also because the changes can be well simulated using appropriate models driven by past emissions changes.

For CO_2, analyses of radiocarbon (carbon-14) changes prove that emissions from fossil-fuel combustion (coal, oil, and gas) have been a major contributor to the concentration increase. Land-use changes (mainly associated with deforestation) have also contributed significantly. For CH_4, the primary sources have been agriculture (rice paddies), animal husbandry, land-fill emissions, and leakage associated with fossil-fuel production and distribution. The main source for N_2O appears to be linked to the use of nitrogen compounds in agriculture as fertilizers. For these three gases, their total emissions are reasonably well defined. Their emissions "budgets" (i.e., the breakdown into different source categories) are more uncertain. The gases do, of course, have important natural sources. However, in pre-industrial times the sources were balanced by natural removal or "sink" processes: by fluxes into the oceans and terrestrial biosphere for CO_2, and, for CH_4 and N_2O, mainly by chemical reactions in the atmosphere. Human activities have disturbed these balances.

For the halocarbons, the most climatically important of which are the chlorofluorocarbons CFC11 and CFC12, the sources are almost all anthropogenic. Today, these sources are largely controlled under the Montreal Protocol and its Amendments and Adjustments. However, new "substitute" chemicals, which are not controlled because they do not cause depletion of stratospheric ozone, are being introduced. These new gases, like all halocarbons, are strong greenhouse gases (although their net effects on future climate are expected to be small relative to CO_2).

In addition to the gases mentioned above, there have been other important atmospheric composition changes due to anthropogenic activities. Emissions of gases like carbon monoxide (CO), nitrogen oxides (NOx), and volatile organic compounds (VOCs) such as butane and propane, which have resulted from industrial activity and land-use changes (biomass burning), have led to large changes in tropospheric ozone. Tropospheric ozone is a powerful greenhouse gas.

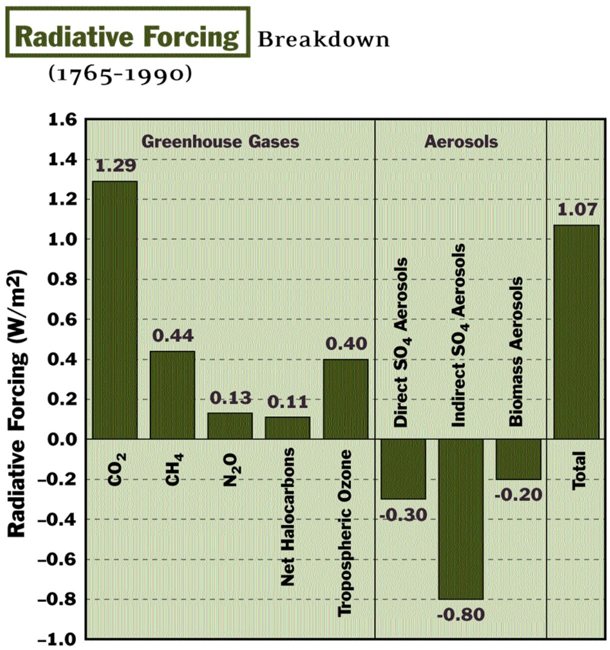
Finally, emissions of SO_2 from fossil-fuel burning (particularly coal), and of other substances released by biomass burning activities, have increased the aerosol loading of the atmosphere. This increase is important because the presence of aerosols has a cooling effect that may partly offset the warming effect of greenhouse gases, as discussed below.

B. Radiative Forcing

The above changes in atmospheric composition have disturbed the overall energy budget of the planet, upsetting the balance between incoming (solar) short-wave radiation and outgoing long-wave radiation — the planet's "radiative balance." Such a change is referred to as "radiative forcing." The climate system responds to positive radiative forcing by trying to restore the radiative balance, which it does by warming the lower atmosphere. The larger the radiative forcing, the larger the eventual surface temperature change.

For each greenhouse gas, and for sulfate and other aerosols, it is possible to calculate the corresponding global-mean radiative forcing. By adding the separate forcings together, we can determine the overall (past or future) external forcing on the climate system. Information on the relationships between forcing and concentration changes (or, for SO_2, emissions changes) has been given by the IPCC (Shine et al., 1990; Harvey et al., 1997). Figure 1 summarizes the forcings over the period from 1765 to 1990. (This figure employs updated forcing relationships for some gases taken from Myhre et al., 1998.)

Figure 1



Breakdown of global-mean anthropogenic radiative forcing from 1765-1990 in watts per square meter (W/m²). These numbers are as used in the IPCC SAR, with updated forcings for CO_2, N_2O and some halocarbons based on Myhre et al. (1998). Methane "CH_4" forcing includes the effect of associated stratospheric water vapor changes. "Net halocarbons" includes the effect of halocarbon-induced stratospheric ozone depletion. "Total" is the sum of greenhouse gas and aerosol forcing.

The numbers in Figure 1 are current best-estimate values. For the greenhouse gases (the first five items in Figure 1), the individual components may be uncertain by up to ±10 percent (Myhre et al., 1998). For total greenhouse-gas forcing, the uncertainty is probably similar. For sulfate aerosol forcing (items 6 and 7) the uncertainty is considerably larger than for greenhouse gases, particularly for the indirect aerosol forcing effect (i.e., the effect these aerosols have on the reflectivity of clouds — see Appendix). The author's judgment, based on a comparison of models and observations (Penner et al., 1997; Wigley et al., 1997), is that the 90 percent confidence interval for total sulfate aerosol forcing from 1765 to 1990 is about -1.1±0.5 W/m².

For the relatively long-lived gases (i.e., gases with characteristic lifetimes of a decade or more: CO_2, CH_4, N_2O, and the CFCs), the spatial patterns of radiative forcing are fairly uniform. For short-lived constituents, which have lifetimes of only days to weeks (e.g.,aerosols and ozone), because their concentration changes are much larger near their sources than elsewhere, the spatial patterns of radiative forcing vary markedly from place to place. Thus, to determine the regional details of past and future climate change, we need to know both the magnitudes and geographical patterns of the emissions that determine historical and future aerosol and ozone concentrations. For the other gases it is sufficient to know only their global emissions changes.

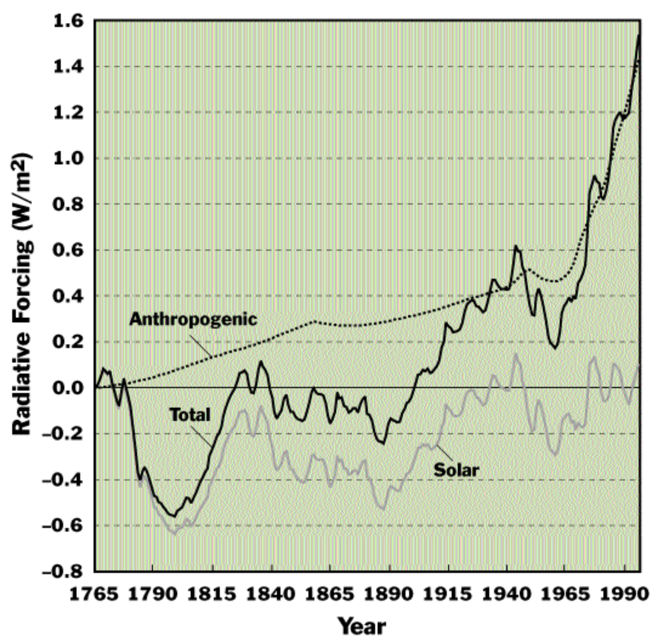
The climate system has experienced more than just anthropogenic forcing since pre-industrial times. In addition, there is strong — but indirect — evidence that appreciable changes have occurred in the energy output of the sun ("solar irradiance"), both on the sunspot cycle (~10 year) timescale and on longer timescales. A number of attempts have been made to reconstruct past changes in the sun's output using sunspot and related data, information from other sun-like stars, etc. (all calibrated against and matched to the satellite-based observational record, which begins only in 1979) (e.g., Hoyt and Schatten, 1993; Lean et al., 1995; Solanki and Fligge, 1998). Prior to the satellite era, even though these reconstructions show qualitatively similar changes, they remain highly uncertain.

Figure 2 compares current estimates for the anthropogenic, solar, and total (anthropogenic plus solar) forcing histories. The solar forcing record used here is that of Hoyt and Schatten (updated; Hoyt, personal communica-

Figure 2

Solar & Anthropogenic Forcing

(1765–2000)



Anthropogenic forcing (greenhouse gases plus aerosols, dotted black line) and solar forcing (gray line) histories, together with their sum (bold line). Since it is change rather than the absolute level that is important, the anthropogenic and solar curves have both been zeroed to the start of the anthropogenic forcing record in 1765. The solar record is from Hoyt and Schatten (1993, updated).

tion). Other records lead to the same conclusions regarding the changing relative importance of solar versus anthropogenic forcing. If we accept the solar record, then we may draw the following conclusions. Until 1890, forcing changes were dominated by solar forcing. From 1890 to 1950, anthropogenic forcing increased by about 0.2 W/m^2, while solar forcing showed a much larger upward trend (0.5 W/m^2). Since 1950, the forcing record is dominated by the anthropogenic component, particularly since 1970. Thus, anthropogenic forcing began to be appreciably larger than natural solar forcing only some 20 to 30 years ago. Before that, natural and anthropogenic forcings were apparently of similar magnitude — indeed, based on the available reconstructions, solar forcing dominated the early part of the record.

C. Changes in Global-Mean Temperature

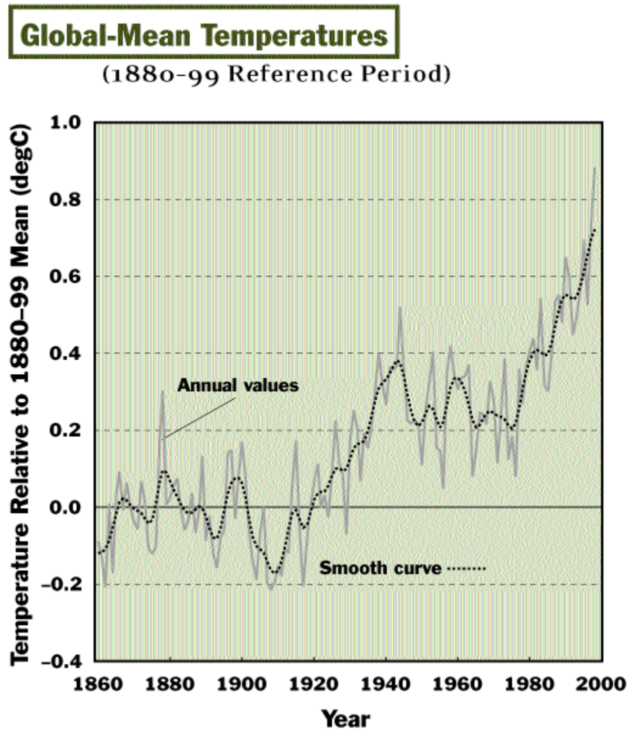
The simplest and most revealing index of climate change is the global-mean temperature near the Earth's surface. Analysis of this record provides us with valuable insights into the causes of past climate change.

The standard record used by the IPCC combines land data developed in the Climatic Research Unit (Jones, 1994) and marine (sea surface) temperature data compiled by the U.K. Hadley Centre (Parker et al., 1995). The raw input data for these records come from many sources, and are subject to numerous inconsistencies arising from nonclimatic effects such as changes in instrumentation, measuring techniques, and the exposure and locations of instruments. (See Jones et al. (1999) for an up-to-date review). Spurious changes may also arise from, for example, urban heat-island effects and coverage changes. Errors arising from these factors have been painstakingly minimized, but small residual uncertainties remain.

The latest record is shown in Figure 3. The most striking feature of this record is the overall warming trend, with the most recent years being the warmest. The record, however, shows a number of other important features. First, there are large variations from year to year. Some of these variations are associated with El Niño, a small number reflect short-term coolings due to volcanic eruptions, and the remainder are probably manifestations of the climate system's own internally generated variability (see Appendix). The record also shows large changes on the 10 to 30 year timescale. These probably reflect anthropogenic and solar forcing effects combined with internal variability.

Critics of the IPCC and the anthropogenic global warming hypothesis often point to the apparent discrepancy between the small greenhouse-gas forcing over 1910–1940 and the rapid global warming that occurred during this period. It is true that this warming was too rapid to be accounted for by anthropogenic forcing alone. However, when the possible effects of internally generated variability and solar forcing are accounted for, there is no serious discrepancy.

Figure 3



Global-mean (land plus marine) temperature changes relative to the 1880-1899 mean as a reference period. The last value shown is 1998. The gray line gives the annual values, while the dotted line gives a smoothed representation to show trends more clearly.

Over the whole period of record, the warming amounts to about 0.6°C since the late 1800s (with a measurement uncertainty of about ±0.1°C). Solar forcing and anthropogenic forcing together are enough to explain the overall warming trend (Santer et al., 1996a; Wigley et al., 1997), although there could be additional influences from factors internal to the climate system (Schlesinger and Ramankutty, 1996). Overall, the observed warming trend is consistent with what we know about the climate system and external forcing changes.

D. Changes in the Free Atmosphere

Human influences on climate are not restricted to the surface. Simple physics demands that any anthropogenic warming should extend throughout the troposphere, primarily because the convective activity associated with clouds keeps this part of the atmosphere well mixed. Above the troposphere, both CO_2 and ozone-depletion effects should have led to cooling, especially in the lower stratosphere. In searching for evidence of human influences, therefore, we need to look not only at the surface record, but also at data above the Earth's surface, in the free atmosphere.

Temperatures above the Earth's surface have been measured since the 1940s. The longest records are those obtained from instruments carried aloft on weather balloons (radiosondes), which are reliable back to the early 1960s. For the troposphere, these data show an overall warming trend, the magnitude of which is very similar to the surface data trend. Over the same period, the data show a marked cooling in the stratosphere. Both the tropospheric warming and the stratospheric cooling are consistent with the predictions of climate models for the joint influences of increasing greenhouse-gas concentrations and halocarbon-induced stratospheric ozone depletion (Santer et al.,1996b; Ramaswamy et al., 1996).

Since 1979, in addition to radiosonde data, a more spatially complete picture is available from space using Microwave Sounding Unit (MSU) instruments on weather satellites (Spencer et al., 1990). Computer weather forecasting models have also been used in recent years to produce syntheses of data from different sources (e.g., Kalnay et al., 1996). In the troposphere, the different records show different trends (Santer et al., 1999). The satellite data show no significant trend, while some radiosonde data show a warming trend that is quite similar to the surface warming trend. In the stratosphere, all records are consistent in showing a marked cooling.

This difference in trends since 1979 between the satellite (MSU) data for the troposphere and the surface data has led some to proclaim that the surface data are flawed and, furthermore, that the lack of a significant MSU trend implies that model predictions of anthropogenic global warming are wrong. Both conclusions oversimplify what is, in fact, a very complex scientific issue. Tropospheric and surface data are different things, so one would not expect them to show identical trends over a period as short as 20 years (Hurrell and Trenberth, 1996). Nevertheless, the differences are large enough to require some additional explanation.

The most obvious explanation for the difference is data uncertainties, which exist for both data sets. For surface data, as noted above, uncertainties arise through instrumentation changes, nonclimatic influences such as urban heat-island effects, and coverage changes and deficiencies. Careful quality control procedures have been applied to minimize these potential error sources.

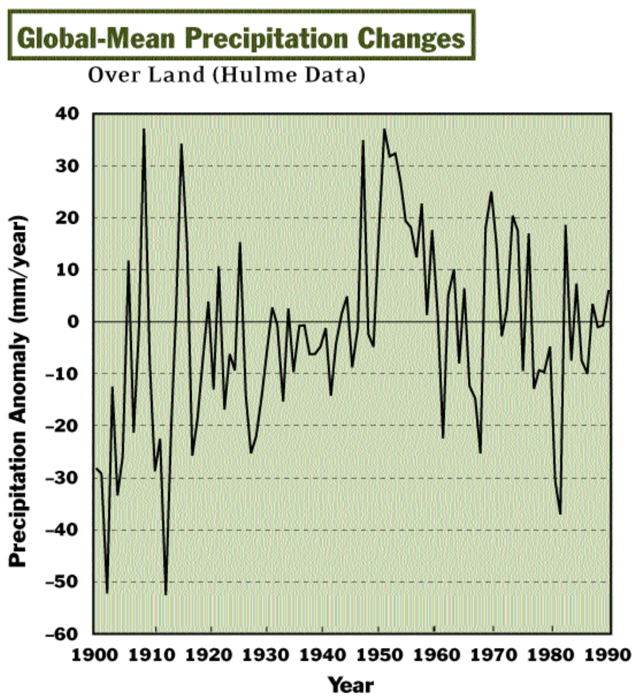
For the satellite data, Hurrell and Trenberth (1998) have suggested that the combination of data from many different satellites required to develop the MSU record has left residual errors in the data, while Wentz and Schabel (1998) have shown that an important correction associated with the satellites' orbital decay was neglected by the satellite data producers. If a correction for orbital decay is applied, the satellite data show a noticeable warming trend and become more consistent with most other data sets (Santer et al., 1999). Christy (1998), however, documents other effects that he claims offset the Wentz and Schabel correction. These satellite data quality issues have yet to be resolved.

An alternative explanation for the satellite-surface trend difference is that it is partially the result of the depletion of ozone in the stratosphere (Hansen et al., 1998). Such depletion would be expected to cool the middle to upper troposphere (a region that is sampled by the satellite record) relative to the surface. This effect, however, does not seem to be enough to fully explain the difference. If stratospheric ozone depletion has affected the satellite record, then, as ozone depletion recovers in response to the Montreal Protocol over the next few decades, tropospheric and surface temperature records should become less divergent.

E. Precipitation Changes

Precipitation is much more variable in both time and space than temperature, and reliable long-term records exist only over the Earth's land areas (e.g., Hulme, 1992; Hulme et al., 1998); and, even here, the coverage is incomplete. Figure 4 shows changes in annual total precipitation averaged over the land areas of the globe (excluding Antarctica) from the Hulme data set. The dominant characteristic of this record is its marked year-to-year variability. If smaller regions are examined, the year-to-year variability becomes even more pronounced. In the assessment of this record in the IPCC Second Assessment Report (SAR) (Nicholls et al., 1996), it is stated (p. 156) that the precipitation data show a small positive (increasing) trend, amounting to +1 percent per 100 years. It can be seen from Figure 4, however, that the apparent trend arises solely because of the

Figure 4



Annual-total precipitation changes averaged over the land areas of the Earth between 55°S and 85°N. These data are from the Hulme (1992) gridded data set. An earlier version of these data was used in the IPCC SAR (Figure 3.11). The values represent anomalies (mm) from the 1961-1990 mean. Since the mean over this reference period was 989 mm, division of the mm value by 10 gives the anomaly in percentage terms.

number of low precipitation years prior to 1915. Unfortunately, one cannot place much confidence in this early part of the record because of data quality problems and reduced spatial coverage. Thus, there is no firm evidence of any real overall trend. There have, however, been some pronounced positive trends in specific regions (see, e.g., Groisman and Legates, 1995; Nicholls et al., 1996).

Because of the high interannual variability (or “noise”) in the precipitation record (see Box 1), associating regional — and/or global — scale precipitation changes with any specific causal mechanism is extremely difficult.

Box 1

Relative Detectability of Temperature and Precipitation Signals

An indicator or index of how easily we might identify (or detect) a human-induced signal in a particular climate record is the signal-to-noise ratio (SNR), i.e., the ratio of the magnitude of the expected anthropogenic signal to the noise level associated with natural variability. It is of interest to compare SNR values for temperature and precipitation.

To be specific, consider the situation prevailing today. Currently, our best estimate of the anthropogenic temperature signal is about 0.4°C (see Figure 5). The noise level of natural variability can be quantified using the standard deviation of the observed temperature record (Figure 3), which is approximately 0.2°C. The current SNR for temperature therefore is two — it would be larger if we were to factor out the human-induced signal and consider only residual variability as noise. A value of this magnitude or higher indicates that the anthropogenic signal should be identifiable in the record.

For global-mean precipitation, the expected signal based on climate model results is about 2 percent per degree of global-mean warming (i.e., if the world warms by 1°C, global-mean precipitation should increase by approximately 2 percent) — see Mitchell et al. (1990), Gates et al. (1992), and Table 2. Since the anthropogenic warming signal is currently about 0.4°C, the corresponding precipitation signal should be about +0.8 percent. Over 1915–1996 (i.e., ignoring the lower quality early years of the record), the standard deviation for precipitation is about 16 mm, or 1.6 percent (see Figure 4). If this value is representative of the global (land plus ocean) noise level, then the implied SNR is 0.5, substantially less than for temperature.

It is, therefore, much more efficient to search for an anthropogenic signal in temperature than in precipitation data.

Apart from changes in average precipitation levels, changes have also been observed in the distribution of precipitation amounts. An important example comes from North America. Here, Karl and colleagues (Karl et al., 1995; Karl and Knight, 1998) have found that the frequency of extreme daily rainfall events (specifically, days with rainfall exceeding two inches) has increased in recent times. They have also shown that the changes are more than one would expect to have occurred by chance. Further, they note that there are qualitative arguments to suggest that similar changes might occur because of greenhouse-gas-induced global warming. These are suggestive results, but they do not prove a cause-effect relationship.

III. Detection and Attribution

The IPCC Second Assessment Report states that "the balance of evidence suggests (that there has been) a discernible human influence on global climate" (Houghton et al., 1996, p. 4). Why did the scientists who wrote the IPCC Second Assessment Report feel able to make such a statement, when, in the previous full IPCC report, they were unable to do so? The critical difference came through the availability of quantitative estimates of the climatic effects of anthropogenically produced sulfate aerosols (Jones et al., 1994; Taylor and Penner, 1994; Boucher and Lohmann, 1995).

Both global-mean and regional-scale data have played important, but complementary roles in recent detection and attribution studies. Global-mean temperature is the variable where we can most easily detect a significant climate change (i.e., one that is highly unusual relative to natural variability). Regional-scale data define the patterns of change, which we need to use to better untangle the individual contributions of human and natural effects in the climate record — and so begin to attribute some part of the observed changes to a human causal factor.

In 1990 (Wigley and Barnett, 1990), it was noted that only the lowest estimates of anthropogenic warming based on model calculations were consistent with the observed changes in global-mean temperature. It was further noted that the pattern of observed temperature change (i.e., the climate change "fingerprint") did not match that expected to arise from increased greenhouse-gas concentrations based on general circulation model (GCM) results.

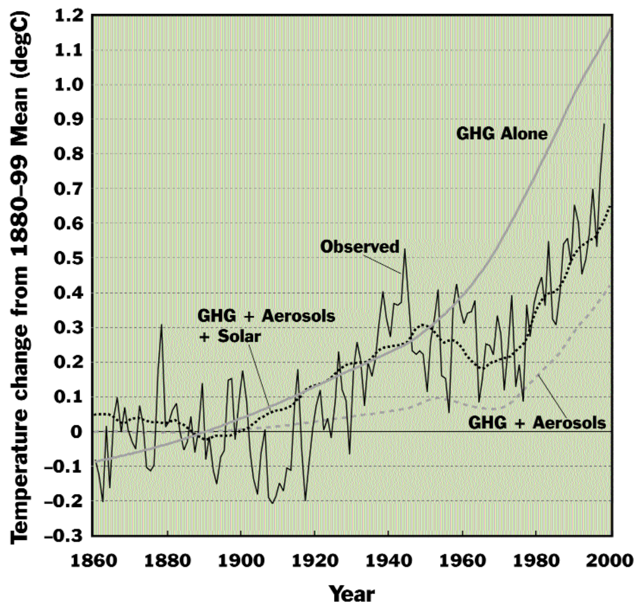
The possibility that sulfate aerosols might account for these discrepancies was first raised in 1989 (Wigley, 1989). At the global-mean level, later calculations (Santer et al., 1996a; Wigley et al., 1997) have shown that the inclusion of aerosol effects can improve the fit between models and observations. If both aerosol effects and the effects of solar forcing are considered, the model-predicted warming is in close agreement with the observations (see Figure 5).

Including the effects of sulfate aerosols has also been shown also to improve the correspondence between model predictions and observed patterns of temperature change, both in the horizontal (latitude-longitude) plane (Santer et al., 1995; Hegerl et al., 1996) and in the vertical (latitude-height) plane (Santer et al., 1996b; Tett et al., 1996). These correspondences, based on rigorous statistical tests, are too close to have occurred by chance. Overall, therefore, there is good agreement between model predictions and observations at both the spatial-mean and spatial pattern levels.

Figure 5

Observed Temperatures

Compared with Model Predictions ($\Delta T2x$ = 2.5degC)



Comparison of observed (thin black line) and model-predicted temperature changes (gray and dotted black lines). The model results use best estimates of greenhouse-gas (GHG) forcing alone (solid gray line), greenhouse-gas plus aerosol forcing (dotted gray line), and greenhouse-gas plus aerosol plus solar forcing (dotted black line). All three model results have been produced using a climate sensitivity of $\Delta T2x = 2.5^{\circ}C$. All changes are shown relative to 1880-1899 as a reference period.

These detection and attribution studies have employed only temperature data. The relative importance of human factors varies greatly according to both the spatial scale and the variable considered. As a general rule, the smaller the spatial scale, the smaller the ratio of human-to-natural influences. Furthermore, the magnitude of the human influence relative to natural variability for temperature is, generally, much larger than for variables like precipitation (see Box 1) and atmospheric circulation (Barnett and Schlesinger, 1987; Santer et al., 1991). These differences are important in understanding future changes. As we continue to perturb the environment with the byproducts of industrial and agricultural activity, so the signal of anthropogenic climate change will continue to grow relative to the background noise of natural variability. For global-mean temperature, anthropogenic warming will become rapidly more and more obvious. However, for other variables like precipitation, and for changes at smaller spatial scales, the human signal will emerge from the background noise much more slowly. In some cases, it may be many decades before we can clearly see these signals.

IV. Predicting Future Climate

A. Future Emissions

The starting point for predicting future changes in climate is usually a “scenario” (i.e., a plausible picture of the future) defining future emissions and/or concentrations of a range of gases. If a scenario involves future emissions, then these must first be translated into future concentrations using appropriate models. The concentrations in turn determine how the balance between incoming short-wave and outgoing long-wave radiation will change; and changes in the radiation balance determine how the climate will change (see Appendix).

It is possible to distinguish two types of emissions scenarios: scenarios that do not explicitly include climate-related policies, and policy scenarios. The former, referred to here as “no-climate-policy” scenarios, give an idea of what might happen in the absence of new policies to limit climate change. Such scenarios are often referred to as “business-as-usual” (BAU) scenarios; but this can be a misleading term, not least because these no-climate-policy scenarios may include the effects of existing or projected policies to reduce other environmental problems such as air pollution and acid precipitation. This is particularly important for SO_2. Only no-climate-policy scenarios are considered here.

Future emissions of the gases that may affect climate depend on future changes in population, economic growth, energy efficiency, and evolving policies to limit emissions. Once these determinants have been specified, they can be used in multidisciplinary integrated assessment models to define future emissions scenarios. Because the determinants are uncertain, a wide range of emissions scenarios can be produced even in the absence of emissions limitations policies. The six no-climate-policy scenarios (IS92a, b, c, d, e, and f) devised by IPCC in 1992 (Leggett et al., 1992) provide an example.

The IS92 scenarios have some well-recognized limitations. For this reason, and because a number of years have passed since they were constructed, a new set of no-climate-policy scenarios is being developed for an IPCC Special Report on Emissions Scenarios (SRES). Preliminary versions of four “marker” scenarios were released by the SRES writing team in December 1998, for use by the international scientific community in climate model simulations that will, in turn, be used in the IPCC Third Assessment Report. These are referred to as the SRES A1, A2, B1, and B2 scenarios.[2] It should be noted that, at the time of this writing, these scenarios have not yet been approved through the formal IPCC review process. They are, however, the most up-to-date and comprehensive emissions scenarios available. The four marker scenarios are used here with permission from the groups and individuals who produced them.

The most marked difference between the SRES scenarios and the earlier IS92 scenarios is in the emissions projections for SO_2. For this gas, the IS92 scenarios did not fully consider the effects of policies to combat air pollution and acid rain (Alcamo et al., 1995, pp. 281, 282). The new SRES emissions scenarios include, in more realistic and internally consistent ways, the possible effects of such policies (Grübler, 1998). In the IS92 scenarios, SO_2 emissions generally increase markedly — e.g., in IS92a from 75 TgS/yr in 1990 to roughly double this in 2050 (1 TgS/yr means 1 teragram, or one million metric tons, of sulfur equivalent per year). In contrast, the new SRES scenarios project eventual decreases in SO_2 emissions over the next century. Since SO_2 emissions lead to the production of sulfate aerosols, which have a strong cooling effect, climate projections based on the SRES scenarios are likely to differ markedly from those based on the IS92 scenarios. Specifically, the reduction in SO_2 emissions and the atmosphere's sulfate aerosol loading will result in increased radiative forcing and warmer temperatures.

B. Future Concentrations and Radiative Forcing

Given an emissions scenario, concentrations may be determined using models that relate changes in atmospheric concentration of a gas to the atmospheric inputs (emissions) and outputs (physical and chemical sink processes). Such models are referred to as gas-cycle models. The predicted concentrations may then be interpreted in terms of their radiative forcing consequences.

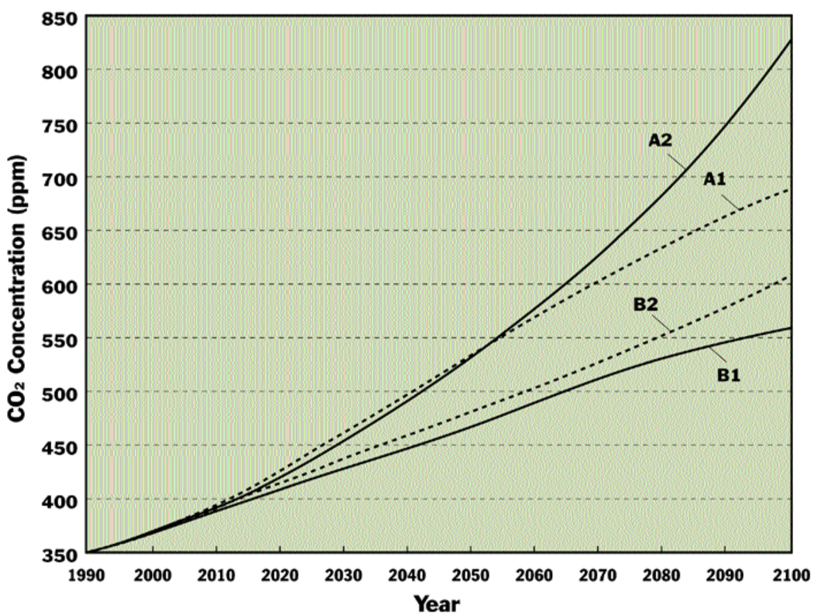
Concentration projections for CO_2 (derived by F. Joos using the Bern carbon cycle model, as used by IPCC; Joos et al., 1996) are shown in Figure 6. The values are similar to those for the IS92 scenarios, but, because the emissions range is smaller, they span a range in 2100 (558 – 825 ppm) that is somewhat narrower than for the IS92 scenarios (488 – 944 ppm). These values are subject to uncertainties arising from uncertainties in our ability to model the carbon cycle (see, e.g., Schimel et al., 1996). In terms of their climate consequences, however, these uncertainty effects are relatively small (±7 percent for global-mean temperature changes from 1990 to 2100 in the assessment of Wigley and Smith, 1998).

Forcing values for the preliminary SRES scenarios are compared with the central IS92 scenario (IS92a) in Table 1. This table clearly shows the dominant role of CO_2 compared with the other greenhouse gases in all scenarios. It also shows how important the new scenarios for SO_2 emissions are. From 1990 to 2100,

Figure 6

CO_2 Concentration Estimates

for Preliminary SRES Scenarios



CO_2 concentration projections for the preliminary IPCC SRES scenarios. Concentrations were calculated with the model of Joos et al. (1996), supplied by F. Joos. They have been adjusted slightly (+1.8 ppm) to match the CO_2 history used in the IPCC SAR.

since SO_2 emissions increase in the IS92a scenario, sulfate (SO_4) aerosol forcing is strongly negative (-0.65 W/m^2), partly offsetting the positive forcing from greenhouse gases during this period. In contrast, since SO_2 emissions decrease from 1990 to 2100 in the more recent SRES scenarios, the corresponding forcing (i.e., relative to 1990) is positive (0.13 – 0.58 W/m^2), adding to greenhouse forcing.

Table 1

Radiative Forcing Estimates

for Preliminary SRES Emissions Scenarios

Component	1765–1990	1990–2050					1990–2100				
		IS92a	A1	A2	B1	B2	IS92a	A1	A2	B1	B2
CO_2	1.29	1.95	2.19	2.17	1.48	1.63	3.69	3.56	4.53	2.44	2.89
SO_4 Aerosol	-1.10	-0.70	0.09	-0.34	0.24	0.17	-0.65	0.58	0.13	0.56	0.19
Other	0.88	0.88	0.83	0.89	0.58	0.97	1.47	0.65	1.82	0.70	0.97
Total	1.07	2.12	3.11	2.70	2.30	2.55	4.51	4.78	6.48	3.70	4.16

Radiative forcing breakdown for the preliminary IPCC/SRES emissions scenarios, compared with the history of forcing over 1765-1990 and with the central IS92 scenario (IS92a). Forcing values were calculated as in the IPCC Second Assessment Report (Kattenberg et al., 1996), except for updating some concentration-forcing relationships following Myhre et al. (1998). "SO_4 aerosol" forcing is the sum of direct (clear-sky) and indirect (cloud albedo) forcing, for which the 1990 values are -0.3 W/m^2 and -0.8 W/m^2. "Other" combines the influences of CH_4, N_2O, halocarbons, tropospheric ozone, stratospheric ozone, stratospheric water vapor, and aerosols from biomass burning.

Total anthropogenic forcing is 1.07 W/m^2 over the 1765 –1990 period. For the future, forcing for the SRES scenarios from 1990 to 2050 ranges from 2.30 to 3.11 W/m^2, in all cases larger than IS92a. All SRES scenarios have CO_2 as the dominant forcing agent, all show important additional forcings due to the sum of other (non-CO_2) greenhouse gases, and all have a positive forcing contribution from sulfate aerosols from 1990 to 2100 (as a result of their SO_2 emissions levels being lower in 2100 than in 1990).

C. Future Global-Mean Climate Projections

In this section, global-mean temperature and sea level projections are given for the preliminary SRES scenarios. These results come from a more comprehensive assessment produced by Smith et al. (in preparation).

The models employed are the same as those used in the IPCC SAR (Kattenberg et al., 1996; Warrick et al., 1996) — see Box 2. To project global-mean temperature changes, the model used is the upwelling-diffusion energy-balance model (UD EBM) of Wigley and Raper (1992; see also Raper et al., 1996). The UD EBM also calculates the amount of expansion of the ocean water mass due to warming. The amount of warming-related melting from glaciers and small ice sheets and from Greenland and Antarctica (Raper et al., 1996; Warrick et al., 1996) is added to this to calculate changes in sea level. The approach used is the same as was used in the IPCC SAR. The only change is in using the preliminary SRES scenarios as the drivers for future change rather than the IS92 scenarios.

Box 2

Simple Climate Models

For global-mean projections, the conventional approach has been to use relatively simple models (the IPCC Second Assessment Report used the upwelling-diffusion energy balance model — UD EBM — of Wigley and Raper, 1992). Such models have both strengths and weaknesses (see, e.g., Harvey et al., 1997). They are less physically realistic than more complex models, since they represent most physical processes in more highly idealized ways. On the other hand, they have the following practical advantages:

(1) They can be run quickly on microcomputers and so can be used to explore the implications of a wide range of emissions scenarios. (This is difficult to do with the most complex climate models because they are highly computationally intensive).

(2) They have user-definable parameters, so they can be used to determine the sensitivity of results to parameter uncertainties (such as those arising from carbon cycle modeling uncertainties, uncertainties in sulfate aerosol forcing, and uncertainties in the climate sensitivity). Complex models like coupled ocean/atmosphere general circulation models (O/AGCMs — see Box 3) have their own specific climate sensitivities.

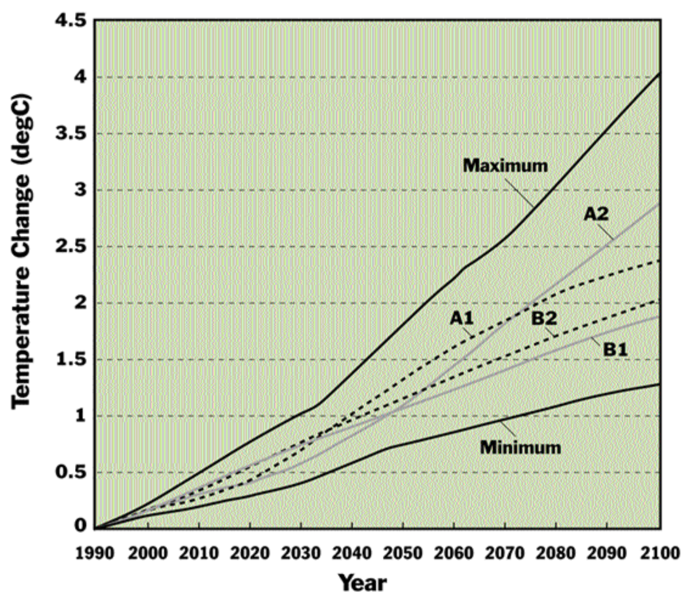
(3) They produce information about the climate change signal directly, unobscured by the “noise” of internally generated climate variability. O/AGCMs have substantial (and generally realistic) internally generated variability, which tends to obscure any underlying signal, particularly when the signal is small.

Coupled ocean/atmosphere general circulation models (O/AGCMs) nevertheless remain the "gold standard" for future climate simulations. Thus, a most important consideration in using simpler models such as UD EBMs is that they should accurately simulate the results of O/AGCMs when used for the same experiments. This was the basis for the use of a UD EBM in the IPCC SAR. Agreement between the simple model used by IPCC and O/AGCM results is demonstrated in Kattenberg et al. (1996), Figures 6.4, 6.13, and 6.17. In essence, simple models are used as relatively sophisticated interpolation and extrapolation tools. They were used in the IPCC SAR to consider a wider range of scenarios than could practically be considered with O/AGCMs, and to assess the magnitude of uncertainties associated with, for example, uncertainties in the climate sensitivity (see, e.g., Kattenberg et al., 1996, Figures 6.20–6.26, and Warrick et al., 1996, Figures 7.6–7.13).

Global-mean temperature and sea level results for the four SRES marker scenarios based on "best-estimate" model parameters are given in Figures 7 and 8. These figures also show the full range of results spanning the scenarios and accounting for uncertainties in the climate sensitivity ($\Delta T2x$) and, for sea level, uncertainties in the ice-melt model parameters.

Figure 7

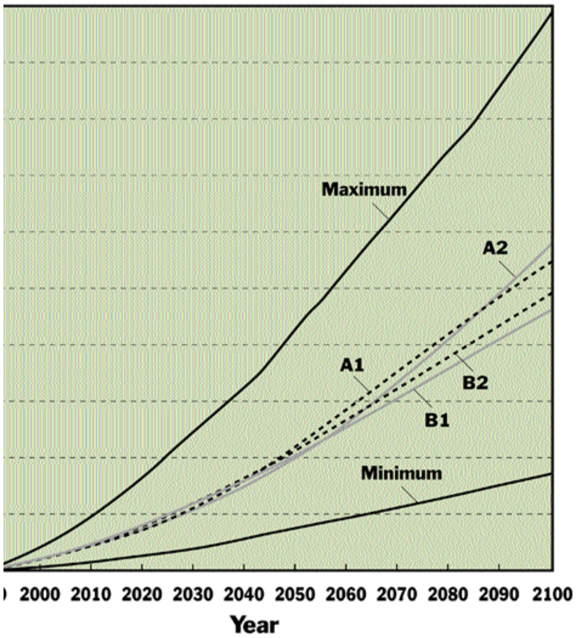
Central **Temperature Estimates** Plus Extremes for the Preliminary SRES Scenarios



The four central curves show best-estimate global-mean temperature changes for the SRES marker scenarios (using a climate sensitivity of $\Delta T2x$ = 2.5°C). The outer curves show extreme values independent of scenario, using sensitivities of 1.5°C or 4.5°C. At different times, different scenarios may represent the extremes. The results shown, therefore, span both the range of scenarios and the range of model parameters.

The central curves in Figure 7 give results for temperature for the four SRES marker scenarios using a climate sensitivity of $\Delta T2x$ = 2.5°C. The global-mean warming from 1990 to 2100 ranges between 1.9°C and 2.9°C. Sea-level rise estimates over the same period for the four scenarios are shown in Figure 8. The individual scenario results for sea level use the temperature estimates from Figure 7 together with central estimates of model parameters used to determine ice melt. The inter-scenario range is 46 to 58 cm. These temperature and sea level results are similar to the central estimates given in the IPCC SAR (namely, those for IS92a with $\Delta T2x$ = 2.5°C) of 2.0°C and 49 cm.

Sea Level Rise Estimates Plus Extremes for the Preliminary SRES Scenarios



The four central curves show best-estimate global-mean sea level changes for the SRES marker scenarios (using a climate sensitivity of $\Delta T2x = 2.5^{o}C$). The outer curves show extreme values independent of scenario, using sensitivities of $1.5^{o}C$ or $4.5^{o}C$ and low and high ice-melt model parameter values. At different times, different scenarios may represent the extremes. The results shown, therefore, span both the range of scenarios and the range of model parameters.

The uncertainty ranges, shown as the lowest and highest curves in Figures 7 and 8, are derived by using the full ranges of climate sensitivity values (1.5 – 4.5°C), ice-melt parameter values, and emissions scenarios. From 1990 to 2100, the range of global-mean warming estimates is 1.3 – 4.0°C. Global-mean sea-level rise over the same period is between 17 cm and 99 cm. The corresponding IPCC SAR ranges (for the IS92 scenarios) are 0.8 – 3.5°C and 13 – 94 cm. The values here are shifted up from those in the IPCC SAR because of the lower SO_2 emissions in the SRES scenarios.

An important point to note is that (as with the IS92 scenarios) the uncertainty range for the SRES scenarios is determined more by climate sensitivity and sea-level modeling uncertainties than by emissions uncertainties, especially for sea level. This is clear if one compares the full uncertainty ranges in Figures 7 and 8 with the spread of results associated with inter-scenario differences. For example, for temperature in 2050, the inter-scenario range is 0.3°C. When climate sensitivity uncertainties are accounted for, the range expands to more than 1.0°C. The differential is even larger for sea level rise (Figure 8).

The results shown in Figures 7 and 8 are a straightforward application of the tools that were employed in the IPCC SAR. Results for the SRES scenarios that will appear in the IPCC Third Assessment Report (TAR) will not necessarily be the same, however, because the TAR will include both results from simple models (such as those used here) and results from a range of coupled O/AGCMs. Furthermore, even simple model results in the TAR will differ from those given here since they will incorporate new scientific knowledge and understanding that has accrued since 1995. The present results should be viewed as a bridge between the IPCC SAR and the TAR, using new emissions scenarios but not applying new modeling “technology.”

V. Regional Climate Change for the United States

The previous section gave a broad (global-mean) picture of the likely magnitude of future climate change. To assess the importance of these changes to the United States and to plan adaptive strategies to minimize potential damages, we need to have information about the spatial details of climate change and their associated uncertainties. This information can be obtained only by using computationally demanding GCMs of the climate system (see Box 3); and, even here, the spatial resolution of such models is quite coarse — 200 km at best.

The ideal tool to use for estimating the spatial details of future climate is the coupled ocean/atmosphere GCM (O/AGCM). A number of simulations of future climate have been carried out with this type of model (e.g., Haywood et al., 1997; Mitchell and Johns, 1997), but, to date, no work has been published in which the simulations use up-to-date combinations of future greenhouse gas and SO_2

Box 3

General Circulation Models

General circulation models represent the Earth's climate system at a discrete series of points, usually with a resolution of a few hundred kilometers in the horizontal plane, and a few kilometers in the vertical. GCMs come in two main types. Both types have a full, three-dimensional atmospheric GCM (AGCM) linked to land-surface and sea ice components. The alternatives then are to couple these components either to a simple "slab" or "mixed-layer" ocean (MLO) model (giving an MLO/AGCM) or to a full three-dimensional ocean GCM (to give a coupled O/AGCM). MLO/AGCMs can only be used for so-called equilibrium experiments in which the eventual changes in climate corresponding to some new (i.e., perturbed) atmospheric composition are determined. O/AGCMs are necessary if one wishes to simulate the time-dependent (or transient) response to a specified time-varying scenario of future atmospheric composition changes.

O/AGCMs, however, are not the only tool that can be used to determine transient climate changes. In Section IV such changes were determined using a UD EBM, but only at the global-mean level. The common factor here is that both types of model have ocean components that characterize the effects of oceanic thermal inertia. An MLO/AGCM cannot do this because it only has a shallow ocean (usually of 50 – 100 meters in depth), whose thermal inertia is very much less than that of the real ocean. Nevertheless, equilibrium MLO/AGCM results can, in certain circumstances, be combined with UD EBM transient results to devise plausible time-varying scenarios with the full spatial detail of an O/AGCM (see Box 4).

A typical example of an equilibrium experiment with an MLO/AGCM is one in which the amount of CO_2 is instantaneously doubled and the model is allowed to come to a new stable (or equilibrium) climate. The modeled climate change is then what would be expected to occur after a considerable period of time had elapsed. In the model, the new state is achieved quite rapidly (because the mixed layer of the ocean has little thermal inertia). In the real world, if we were to suddenly double the CO_2 concentration, it would take decades to centuries to reach a new stable climate because of interactions between the mixed layer and deeper parts of the ocean. Both UD EBMs and coupled O/AGCMs account for this thermal inertia effect, so the time-dependent nature of their response is more similar to what is expected in the real world.

emissions. Such simulations, based on the IPCC SRES scenarios, are currently being carried out by a number of GCM modeling groups for input into the Third Assessment Report.

In the absence of these results, however, we can still gain useful insights into the future by using currently available model results. We do so by comparing and synthesizing these results to provide both an overall assessment of them and an overview of what they predict for regional-scale changes in temperature and precipitation over the United States.

A. Model Evaluation

How credible are currently available GCMs? There are two ways to answer this question. The first is a standard model evaluation procedure: one simply compares the model's simulation of current climate with observations (see, e.g., Gates et al., 1999). Analyses like these give widely varying results. Some models are good in one region and less good in another, and some models perform well for some variables but relatively poorly for others. A second approach is to compare the results of different models when they are all used to perform the same type of climate-change experiment.

For the present analysis, results from 15 different models are compared. The models considered are those compiled in the SCENGEN (climate SCENario GENerator) software package (Hulme et al., 1995). These models have different vertical and horizontal resolutions and represent different model "vintages." Most of the models are MLO/AGCMs (see Box 3), but four are coupled O/AGCMs. Table 2 lists these models, together with the year the particular experiment used here was performed, the model's horizontal and vertical resolution, the primary source reference, the model's climate sensitivity, and its precipitation sensitivity (percentage global-mean precipitation increase per 1°C global-mean warming).

The first part of the present model evaluation is to compare model simulations of present-day climate with observations. Only a single (but quite demanding) criterion is used, the average (over 12 months) of the global pattern correlation between modeled and observed precipitation (Table 2). High values of this correlation indicate that modeled and observed precipitation patterns are similar, and low values point to important differences. A correlation of 0.707 is required for modeled and observed patterns to have 50 percent of their spatial variability in common. Only four models reach this threshold. If one plots this pattern correlation against model year, there is an upward trend pointing to improvements in the models over time. The best model (by this criterion) is the U.K. Hadley Centre's coupled O/AGCM (HadCM2), which has a pattern correlation value of 0.77.[3]

Table 2

General Circulation Models

Model	Experiment Year	Horizontal Resolution (lat. x long.)	No. of Levels	Source Reference	ΔT2x (°C)	Precipitation Sensitivity (% / °C)	Precipitation Pattern Correlation
BMRC	1991	3.2° x 5.6°	9	Colman and McAvaney (1995)	2.2	1.4	0.61
CCC	1989	3.75° x 3.75°	10	Boer et al. (1992)	3.5	1.1	0.63
CSIRO1	1991	3.2° x 5.6°	9	McGregor et al. (1993)	4.8	2.1	0.64
CSIRO2	1995	3.2° x 5.6°	9	Watterson et al. (1997)	4.3		0.71
ECHAM1*	1989	5.6° x 5.6°	19	Cubasch et al. (1992)	2.6	1.8	0.64
ECHAM3*	1995	5.6° x 5.6°	19	Voss et al. (1998)	2.6		0.67
GFDL	1986	4.5° x 7.5°	9	Wetherald and Manabe (1986)	4.0	2.3	0.58
GISS	1983	8° x 10°	9	Hansen et al. (1984)	4.2	2.8	0.58
HadCM2*	1994	2.5° x 3.75°	11	Mitchell et al. (1995)	2.5		0.77
LLNL	1989	4° x 5°	2	Pollard (1982)	3.8	2.6	0.56
OSU	1988	4° x 5°	2	Schlesinger and Zhao (1989)	2.8	2.9	0.59
UIUC	1996	4° x 5°	11	Schlesinger (1997)	3.4		0.65
UKHI	1989	2.5° x 3.75°	11	Senior and Mitchell (1993)	3.4	2.7	0.72
UKLO	1986	5° x 7.5°	11	Wilson and Mitchell (1987)	5.2	3.0	0.64
UKTR*	1991	2.5° x 3.75°	11	Murphy and Mitchell (1995)	2.7	1.8	0.76

GCMs used in the present study. The horizontal resolution and number of levels refer to the atmospheric components of the models.

**denotes coupled O/AGCM*

The second part of the present model evaluation is to compare the results of different models for a similar climate change experiment. If all models are asked the same question, how well do the models agree? Lack of agreement would imply that there is considerable uncertainty regarding regional-scale climate-change results, and that one should be cautious in using results from any one model. Model agreement, of course, would not guarantee that their results were unequivocally correct.

The experiment used here for inter-model comparison is one where the CO_2 concentration is doubled (or increased progressively in an O/AGCM experiment). For our test of agreement (or otherwise), results over the continental United States only have been used (spanning 27.5–52.5°N, 67.5 –122.5°W). The data used were seasonal-mean changes in temperature and precipitation for winter (December, January, February); spring (March, April, May); summer (June, July, August); and fall (September, October, November).[4]

The comparisons show that some model pairs have very similar patterns of change, while other pairs give highly dissimilar results. The best results (i.e., greatest consistency between models) are obtained for winter temperature-change patterns, largely because many models show an enhanced warming in higher latitudes. The worst results (i.e., greatest inter-model differences) are for summer and fall precipitation-change patterns. Here (and for precipitation in general) inter-model differences are generally very large.

For temperature, the modeled changes are always larger than any differences between the models. In other words, there is a clear warming signal over the whole region and in all seasons that is common to all models. For precipitation, the inter-model comparison results are less satisfactory. Generally, the average signal is smaller than the average difference between the models. This is particularly the case in summer and fall. There is a clearer signal in winter and spring in the northern 10° latitude band of the study area (which is mainly in Canada).

The mean (model average) patterns of temperature change are shown in Figure 9. To more easily compare models, and to make the results applicable to different future times, the results that are averaged are the changes relative to each model's global-mean temperature change. In other words, to interpret the results in Figure 9, one has to multiply them by the global-mean temperature change. For example, for a 2ºC global-mean warming, the values in Figure 9 would have to be doubled to find the corresponding regional and seasonal-mean warming.

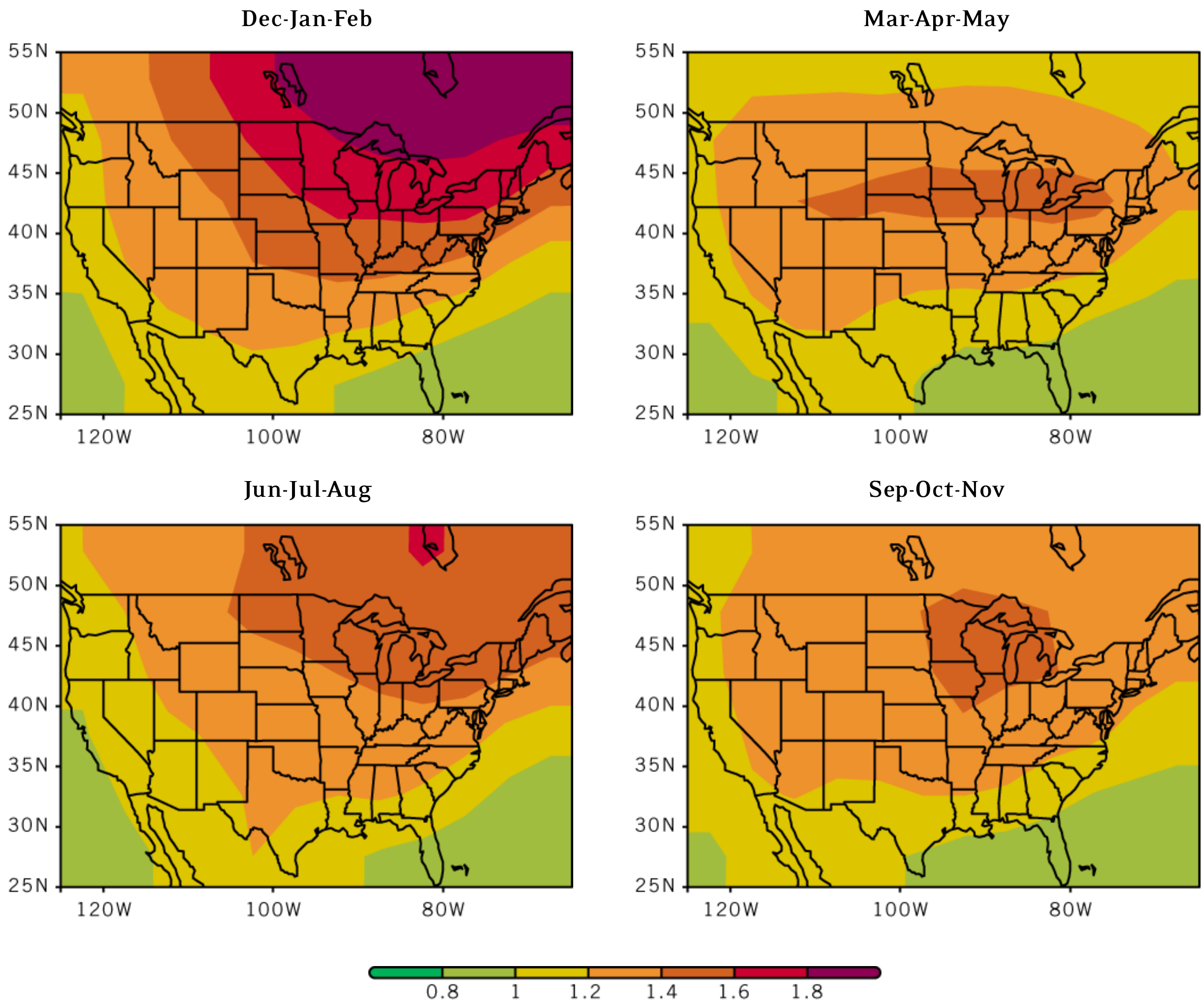
An important conclusion from this figure is that, in almost all parts of the lower 48 states and in all seasons, the warming exceeds the global-mean warming (i.e., the values shown in the figure are greater than one). The Southeast and Southwest are the exceptions in that they tend to show warming slightly below the global mean. At the other extreme, in winter, the northernmost states from North Dakota eastward to Maine show enhanced warming by a factor of up to two relative to the global mean.

Model-average results for precipitation change are shown in Figure 10. Again, these represent changes per 1ºC global-mean warming. Note that while the bulk of the study area shows precipitation increases, the changes (ranging between -4 percent and +8 percent per 1°C global-mean warming) are small everywhere relative to current levels of interannual variability. This is a consequence of the large inter-model differences, which leads to canceling of disparate model results and, hence, to a relatively weak residual signal. Individual models, however, can show very large regional changes. As an example, Figure 11 shows precipitation changes for the HadCM2 model: changes here range from around -10 percent to larger than +20 percent per 1°C global-mean warming. Note that even the small changes shown in Figure 10 could have important consequences for sectors such as agriculture and water resources, as noted in other reports of this Pew Center series (see Adams et al., 1999; Frederick and Gleick, in preparation).

Figure 9

Relative Temperature Changes

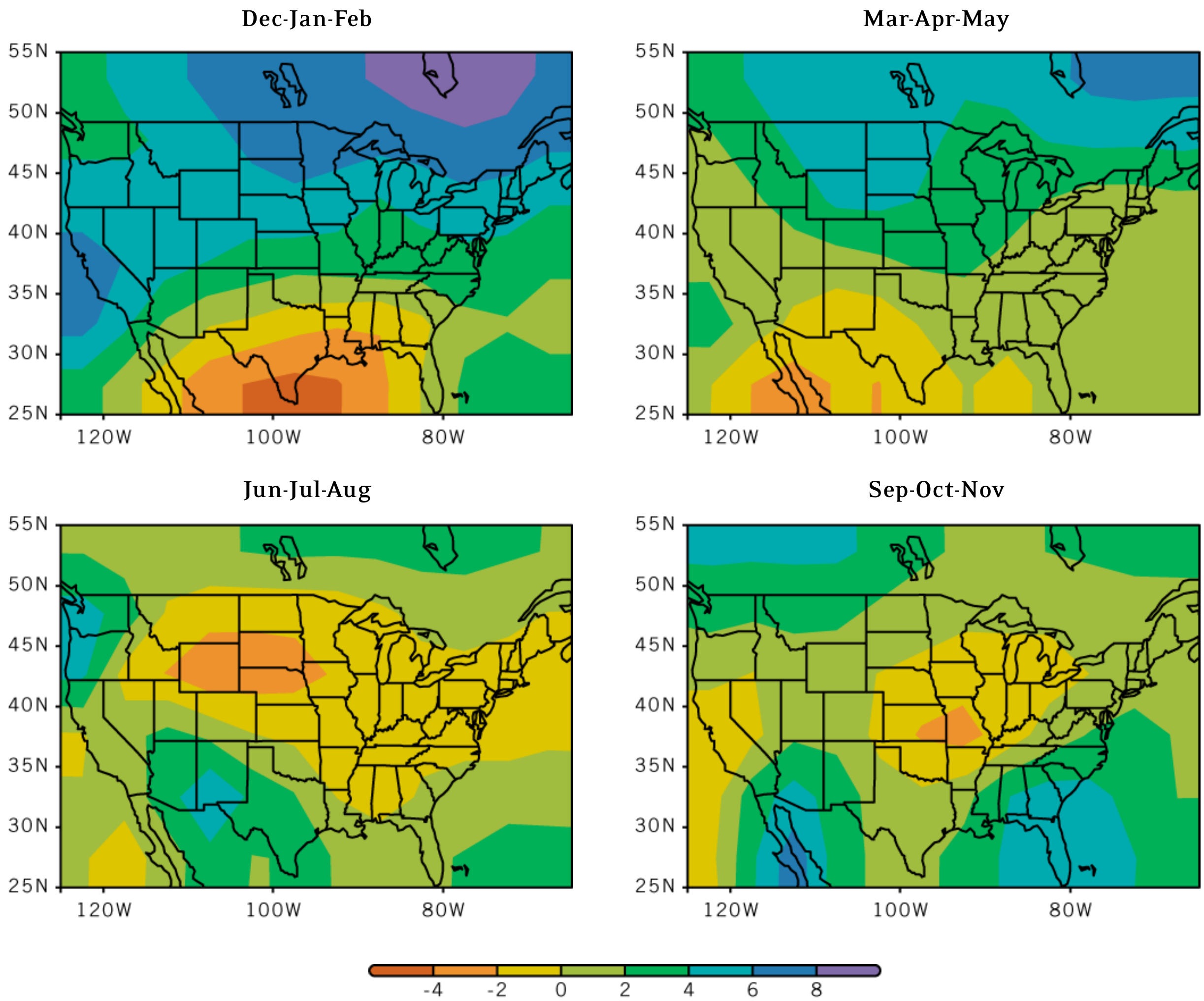
Mean of 15 Models



Ratio of model-average temperature change to change in global-average temperature for the effect of greenhouse gas increases. The four maps show results for the four seasons, winter (December, January, February); spring (March, April, May); summer (June, July, August); and fall (September, October, November). These results may be used to estimate changes at any future date simply by taking the global-mean temperature change for that year (e.g., from Figure 7) and using this to multiply the values shown in the maps.

Figure 10

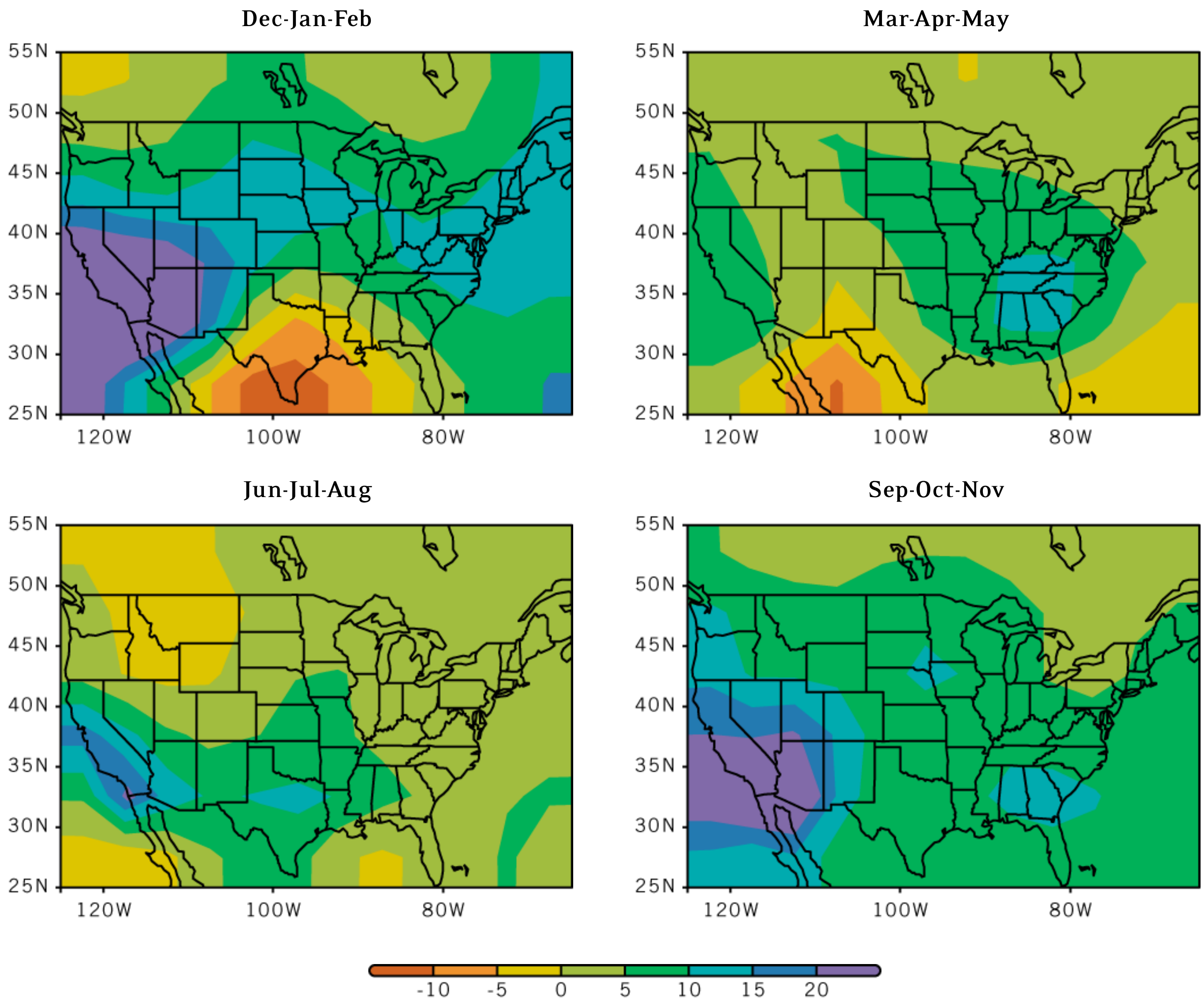
Relative Precipitation Changes (%/°C) Mean of 15 Models



Model-average precipitation changes (percent) relative to the global-annual-mean temperature change for the effect of greenhouse gas increases (i.e., percent change per 1 °C global-mean warming). The four maps show results for the four seasons, winter (December, January, February); spring (March, April, May); summer (June, July, August); and fall (September, October, November). These results may be used to estimate changes at any future date simply by taking the global-mean temperature change for that year (e.g., from Figure 7) and using this to multiply the values shown in the maps.

Figure 11

Relative Precipitation Changes for HadCM2 (%/°C)



Precipitation changes (percent) for the U.K. Hadley Centre's coupled O/AGCM (HadCM2) expressed relative to the global-annual-mean temperature change for the effect of greenhouse gas increases (i.e., percent change per degree Celsius). The four maps show results for the four seasons, winter (December, January, February); spring (March, April, May); summer (June, July, August); and fall (September, October, November). These results may be used to estimate changes at any future date simply by taking the global-mean temperature change for that year (e.g., from Figure 7) and using this to multiply the values shown in the maps.

B. Future Climate

Even though the results presented in the previous section do not include the effects of sulfate aerosols, they can still provide useful information about future climate change possibilities over the United States. The method for doing this is described in Box 4, and can be illustrated with a specific example.

Suppose we are interested in the implications of the SRES A1 scenario and that we wish to derive patterns of climate change for a period centered on 2030. Suppose further that the climate sensitivity is assumed to be $\Delta T2x = 2.5°C$ (the current IPCC best-estimate value). To obtain patterns of climate change for 2030, one simply reads the global-mean warming directly from Figure 7 (namely, 0.7°C) and scales the normalized patterns of change (Figures 9 – 11) by 0.7. To obtain an absolute climate scenario, one would add these changes to the current (1990) climate.

Sulfate aerosol effects will undoubtedly modify these results. At the global-mean level, the forcing contribution from sulfate aerosols is small relative to the total forcing (generally less than 15 percent, see Table 1). However, because of the large spatial variability in the emissions of SO_2 and the forcing from sulfate aerosols, there may still be important effects at the regional level. These effects will vary with emissions scenario and time. At present, it is not possible to give any reliable indication of what they may be, partly because appropriate O/AGCM model experiments have yet to be performed, but also because of the very large uncertainties surrounding the quantification of the relationships between SO_2 emissions and the resulting forcing effects.

Box 4

Climate Scenario Construction

A scaling method developed by Santer et al. (1990) allows us to use results from any model, no matter what its climate sensitivity, to devise climate change scenarios rapidly for arbitrary emissions scenarios and to assess the effects of a wide range of sources of uncertainty. The procedure is as follows. First, the patterns of climate change from either an MLO/AGCM or an O/AGCM are scaled by the model's global-mean warming to produce "normalized" patterns of change, i.e., patterns of climate change per 1°C global-mean warming. For example, if the global-mean warming for a specific model were 3.0°C, then we would divide all of the changes produced by that model, gridpoint by gridpoint, by 3.0. This is a useful unifying procedure, since it makes results from different models more directly comparable by removing differences associated with their different climate sensitivities.

The next step is to scale these normalized patterns up or down by whatever the best estimate of global-mean warming happens to be — or by a range of estimates if we want to explore uncertainty issues. The warming value used for scaling will depend on the emissions scenario being considered, the selected value for the climate sensitivity, and the future point in time that is of interest.

VI. Changes in Other Aspects of Climate

The previous sections have stressed the uncertainties that surround projections of future climate change both at the global-mean level and, even more so, at the regional level. For the latter, only temperature and precipitation were considered. Over the United States, one can be fairly confident that the warming will be greater than the global-mean warming worldwide, with greatest enhancement at high latitudes in winter.

For precipitation, the changes are far more uncertain, largely because different models give widely differing results. The only result common to most models is a precipitation increase in winter over the northern Great Plains/Great Lakes region, and northeastern states. In the central and southern latitude bands of the United States, some models show substantial increases in precipitation, while others predict substantial decreases.

The impacts of climate change at any particular location will, however, be determined by factors other than just changes in mean temperature and precipitation. A summary of what is known about some of these other factors is presented below.

A. Extremes of Temperatures

A general warming will shift the whole distribution of temperatures. Thus, relative to any fixed threshold, the frequency of warm temperature extremes (on all timescales — days, seasons, and years) will increase and the frequency of cold extremes (like frost days) will decrease. This is a general result, applicable to any part of the globe. In the absence of variability changes, the increase in the frequency of extreme warm events will be disproportionally large (Wigley, 1985). For example, if a 1°C warming increased the number of days over a particular threshold by 10 percent, then a 2°C warming would cause an increase by substantially more than 20 percent.

B. Changes in Variability

Changes in variability are important because they may have a significant effect on agriculture and water resources (see, e.g., Mearns et al., 1996). Furthermore, the IPCC Second Assessment Report (Houghton et al., 1996, p. 44) notes that "a small change in variability has a stronger effect (on the frequency of extremes) than a small change in the mean," as pointed out earlier by Wigley (1985).

There is, however, no consensus between models on changes in the interannual variability of climate elements like temperature and precipitation. Indeed, even the best models (such as HadCM2) perform poorly in simulating such variability (e.g., Tett et al., 1998) — i.e., their simulations of current variability differ noticeably from observed variability. If any changes did occur, they would be regionally specific, so that some regions might experience increases in variability while nearby regions might experience changes in the other direction.

C. Changes in Precipitation Extremes

The IPCC Second Assessment Report (Houghton et al., 1996, p. 44) notes that GCM results suggest increases both in the frequency of intense precipitation events and, in some regions, in the probability of dry days and the length of dry spells (see, e.g., Fowler and Hennessy, 1995; Gregory and Mitchell, 1995). Two more recent studies support this conclusion. Zwiers and Kharin (1998) found that heavy precipitation events over North America might occur twice as often in a world that was 3.5°C warmer than today. Frei et al. (1998) found a similar shift to more frequent heavy precipitation events in southern Europe.

While these analyses are careful and comprehensive, one must still be cautious in accepting their quantitative conclusions. In both studies, the warming considered is substantially greater than that expected over the next 50 years. As an additional cautionary note, Osborn (1997) has shown that one cannot automatically translate changes in precipitation intensity at the GCM gridbox level to real-world local changes. In some cases, in making this spatial-scale conversion, an increase in intensity can become a decrease (or vice versa).

Thus, while both types of change (more frequent wet extremes and dry extremes) are possible in the United States, there is no unequivocal evidence for either. Furthermore, the large inter-model differences in projections of mean precipitation change shown elsewhere in this report imply that one should treat the predictions of single models cautiously, especially for changes in the shorter time-scale events referred to above.

While we can say little about precipitation changes in most parts of the United States (except for the increase in precipitation in higher latitude regions in winter), a general statement can be made about the overall hydrologic budget. Since warming should lead to increased evaporation, if precipitation were not to change at all at a particular location, soil moisture levels and the availability of water for runoff would have to decrease (see, e.g., Manabe et al., 1981; Wigley and Jones, 1985). However, even this conclusion is subject to uncertainty because of the direct plant-physiological effect of increasing CO_2 concentrations on plant water-use efficiency. If, as small-scale experiments suggest, water-use efficiency increases with increasing CO_2, then plants would transpire less in the future. To some degree, at least, this would offset any tendency toward increased evaporation as a result of warming. The big uncertainties here are in scaling up the small-scale experimental results to larger, ecosystem scales, and in knowing how ecosystems will respond to future time-varying changes in climate.

D. Midlatitude Storms

For midlatitude storm systems, the state of science is exemplified by IPCC's cautious statement that ". . . there is little agreement between models on . . . changes in storminess . . . (and) conclusions regarding extreme events are obviously even more uncertain" (Houghton et al., 1996, p. 44).

E. Hurricanes and Tropical Storms

The formation of tropical storms is controlled by many different factors, including sea surface temperatures, atmospheric stability, wind shear (i.e., wind direction changes with height), the large-scale circulation in which a storm may be embedded, and high-level wind patterns (see e.g., Henderson-Sellers et al., 1998). Current GCMs used in climate studies do not have fine enough spatial resolution to be able to simulate individual tropical cyclones (i.e., in the jargon of the field, such storms are sub-gridscale events). Furthermore, even the most sophisticated weather forecasting models are generally unable to predict the initiation of tropical cyclones. Not surprisingly, therefore, our knowledge of how climate change might affect the frequency, intensity, or tracks of tropical cyclones is highly uncertain.

Nevertheless, there is empirical evidence that there might be small increases in the frequency of Atlantic hurricanes (Raper, 1993), based on the positive correlation between SSTs and hurricane frequencies in this region. (Such correlations are much weaker in other regions; in some areas they are negative.) There is also model evidence that minimum pressures may decrease and windspeeds may increase in tropical storms worldwide. Knutson et al. (1998), for example, project windspeed increases of 5 to 12 percent for a sea-surface temperature increase of 2.2°C (a rise that might occur by 2100). However, the projected changes are small relative to past interannual variability. Thus, even if these projections could be considered reliable, it would be many decades before the hypothesized signals could be positively detected above the noise of interannual variability.

An associated possibility is that, along with a minor intensity increase, there could be substantially larger changes in the amount of precipitation associated with individual storms (Knutson and Tuleya, 1999). This may be a more robust result because, with increased ocean temperatures, it is almost certain that the moisture-holding capacity of the atmosphere will increase. Along with this, one would expect increased precipitation at the global-mean level. While the manifestation of this general increase over midlatitude land areas is highly uncertain, more confidence can be placed on the possibility of precipitation increases in areas currently frequented by tropical cyclones (and, as noted previously, in higher latitudes).

VII. Conclusions

Since the late 1800s, both the atmospheric concentrations of greenhouse gases (CO_2, CH_4, N_2O, etc.) and the atmospheric loading of sulfate aerosols have increased markedly, due almost entirely to human activities. At the same time, the average surface temperature of the Earth has warmed by about 0.6°C. There is strong evidence that the two are related: indeed, when the radiative forcing effects of greenhouse gas and aerosol increases are considered together with those due to estimated changes in solar output, agreement between model predictions of global-mean temperature and observed changes is excellent. Furthermore, the observed patterns of temperature change, both at the surface and in the zonal-mean/vertical plane, also agree well with model predictions of anthropogenic forcing effects. It is highly unlikely that such agreements could have occurred by chance or be due to natural climatic variability. Such consistency between observations and model expectations at the global-mean and spatial-pattern levels should engender confidence in the models and, therefore, in the broad-scale features of their projections.

This report has developed new estimates of changes in global-mean temperature and sea level rise using (with permission from the producers) preliminary versions of four new marker emissions scenarios (the SRES scenarios). If central estimates of model parameters are used, global-mean warming from 1990 to 2100 ranges from 1.9°C to 2.9°C. Sea-level rise estimates over the same period range from 46 to 58 cm. The ranges here arise solely from differences in the emissions scenarios. For temperature, these values represent warming rates between three and five times the rate of warming that has occurred over the past century. When the full range of emissions, climate sensitivity and ice-melt model parameters is considered, the global-mean temperature change from 1990 to 2100 ranges between 1.3°C and 4.0°C while the sea-level rise ranges between 17 cm and 99 cm.

Regional changes may differ markedly from global-mean changes. For the United States, based on results from a number of climate models, the rate of future warming is expected to be noticeably faster than the global-mean rate. Future regional-scale precipitation changes are highly uncertain. The only result that is common to all climate models is an increase in winter precipitation in northern latitudes, from the northern Great Plains to the northeastern states.

Changes in weather and climate extremes over the United States are certain to occur as the global climate changes. For hurricanes and tropical storms, model-based evidence suggests that there could be small increases in their intensity: i.e., lower central pressures and higher windspeeds. Further, such storms will probably be accompanied by larger rainfall amounts. Empirical evidence suggests that a small increase in frequency of hurricanes is possible in the North Atlantic region. The frequency of extremely hot days is almost certain to increase, and the frequency of frosts should decrease. Changes in the frequency of daily precipitation extremes are highly uncertain, although there is evidence for an increase in the frequency of wet extremes. For all extreme events, however, it is unlikely that the projected changes will become evident in a statistically convincing way for many decades, with the exception of temperature extremes, which should become evident sooner.

Endnotes

1. In the United Nations Framework Convention on Climate Change (UNFCCC), the term "climate change" is used to refer only to human-induced (or "anthropogenic") change. The scientific usage of this term is more general, referring to all climate change, both natural and anthropogenic.

2. The SRES marker scenarios are based on a set of "storylines" that define parameters such as future population levels, economic growth rates, energy technologies, levels of international cooperation, etc. The storylines and their background are described on the web site http://sres.ciesin.org/sres/htmls/storyline.html. In brief, A1 is a future world with rapid economic growth and rapid introduction of new and more efficient technology, low population growth, and a substantial reduction in regional differences in per capita income; A2 corresponds to a very heterogeneous world with high population growth and less concern for rapid economic and technological development; B1 has low population growth, embraces rapid changes in economic structures, a move toward a less materialistic society, the introduction of clean technologies, rapid technology development, and an emphasis on global solutions; B2 has moderate population growth, less rapid but more diverse technological change, and an emphasis on local solutions and environmental sustainability. The four marker emissions scenarios were produced by the following groups: A1, Centre for Global Environmental Research, National Institute for Environmental Studies, Tsukuba, Japan, using the Asia-Pacific Integrated Model (AIM), Morita et al. (1998); A2, ICF Consulting Group, Washington, D.C., in collaboration with the U.S. EPA, using the ASF model structure, Pepper et al. (1998); B1, RIVM, The Netherlands, using IMAGE 2.1, Alcamo et al. (1998); B2, IIASA, Laxenburg, Austria, using the IIASA model, Nakićenović (1999). Further documentation is to be published in special issues of the journals *Mitigation and Adaptation Strategies for Global Change* and *Technological Forecasting and Social Change*. Anthropogenic emissions results for 2100 for CO_2 and SO_2 are shown below, compared with 1990 values.

	1990	A1	A2	B1	B2
Fossil CO_2 (GtC/yr)	6.2	13.2	28.8	6.5	13.7
Net deforestation (GtC/yr)	1.1	-0.6	0.2	1.4	-0.2
SO_2 (TgS/yr)	72	28	61	29	47

3. While this may not seem high, it should be noted that even the observed data are subject to significant uncertainties. Different data sets purporting to show the same thing do not correlate perfectly. A typical interdata-set correlation is about 0.9 (Hulme, 1992), so a model-observed correlation of 0.77 is quite good relative to this. Furthermore, simulating observed precipitation patterns is perhaps one of the most stringent tests that can be applied to a model, not only because precipitation is such a highly variable quantity, both in time and space, but also because it is the result of a great many complex physical processes, many of which operate on spatial scales well below the smallest scale that a GCM can resolve.

4. Three methods of comparison were employed: the use of pattern correlations to compare climate-change patterns between models; a comparison of the average pattern (averaged over all models) with a measure of inter-model differences; and a simple count of the number of models giving changes in the same direction. The full results are not given here — they may be obtained from the author in a supplementary document to the present report.

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Appendix

A1 The Climate System

Predicting future climate is a daunting task. To do so we must consider not just the atmosphere, but also the oceans, the cryosphere (i.e., land-based and marine snow and ice), the land surface, the stratosphere, and the sun, together with the interactions between these different components. These various components of the climate system involve processes that act on a vast range of spatial and temporal scales.

Because of these complexities, the only practical approach to climate prediction is to use mathematical models of the various processes and interactions, and to run these models on computers. There is a hierarchy of models that may be used, depending on the degree of detail required in the prediction. For example, if we wish only to estimate how the global-mean temperature might change, we can use a relatively simple model that can be run on a personal computer. If we wish, however, to estimate how temperature, rainfall, storminess, and other aspects of the weather might change at a particular place, then we need to use a much more sophisticated model (similar to, but even more complex than a weather forecasting model). These models, called general circulation models (GCMs), when used for extended (multi-decadal) climate simulations, can be run only on the world's most powerful computers.

A2 External Forcing and the Earth's Energy Balance

The causes of changes in climate can be divided into those due to external forcing and those that occur because of factors internal to the climate system (referred to as natural internal variability). External forcing effects can, in turn, be divided into natural influences and human influences.

The temperature of the Earth (i.e., the global-mean, near-surface temperature) is determined by the balance between incoming solar energy (which is in the short-wave region of the electromagnetic spectrum) and outgoing energy, which comprises both reflected (short-wave) solar energy and long-wave energy radiated back into space from the Earth's surface and the atmosphere. The amount of outgoing long-wave energy depends on the temperature of the Earth's surface, the temperatures in the atmosphere above the surface, and the distributions of gases and clouds. The effect of these together may be represented crudely as an "effective planetary temperature," the value of which can be quite different from the actual surface temperature.

Normally, on timescales of decades and longer, there is a balance between the amount of incoming and outgoing energy. This means that the effective planetary temperature remains approximately constant on these timescales. If the balance is disturbed, the imbalance is referred to as "radiative forcing." The planet responds to radiative forcing by attempting to restore the radiative balance, which it does by changing the effective planetary temperature.

Processes that lead to an imbalance (i.e., to radiative forcing) are generally called "external forcing." An example of external forcing would be a change in the energy output of the sun. If the sun's output were to wax and wane, the Earth's effective planetary temperature would move up and down in response to maintain the radiative balance. This does not mean, however, that the temperature at the Earth's surface will change in exactly the same way. Within the Earth-atmosphere system, actual temperatures vary widely, and the relationship between the effective planetary temperature and actual temperatures depends on many factors.

The Earth's effective planetary temperature depends on the amount of energy received from the sun and on the composition of the atmosphere; the amount and distribution of clouds and the concentrations of gases that can absorb long-wave energy, the so-called greenhouse gases - water vapor, CO_2, CH_4, N_2O, etc. The fact that these greenhouse gases exist naturally causes the near-surface layers of the atmosphere to be warmer than they would otherwise be. Since this effect is superficially similar to the way a plastic or glass greenhouse traps heat within its walls, we refer to it as the "greenhouse effect." If a greenhouse gas increases in concentration, this leads to further near-surface warming - the "enhanced greenhouse effect." The enhanced greenhouse effect is another example of external forcing.

The near-surface temperature can also be changed by changing other components of the energy balance, either the amount of incoming solar radiation (already noted) or the amount of outgoing, reflected short-wave radiation. Both are additional examples of external forcing. An example of a change in reflected radiation might arise if we were to directly change the character of the Earth's surface (e.g., by deforestation). This would change the reflectivity (or "albedo") of the surface because of the different reflectivities of different vegetation types.

Note that changes in surface reflectivity might also occur as a result of climate change. For example, global warming will most likely reduce high-latitude snow and ice cover and so reduce the planet's overall albedo, which in turn would change the forcing balance and lead to additional warming. This is an example of a positive "feedback" mechanism, not of directly imposed external forcing. A related example occurs when an externally forced climate change leads to changes in clouds (cloudiness, cloud types, cloud levels, etc.). Cloud changes affect both the amount of outgoing long-wave radiation and the overall reflectivity of the planet, and so provide another important feedback mechanism.

Another important example of external forcing is provided by the effects of small particles in the atmosphere, referred to as aerosols. There are many different types of aerosol, including mineral dust, salt, soot, organic material, and sulfate aerosols. These are produced both naturally and as a result of human influences. Anthropogenic aerosols include sulfate (SO_4) aerosols produced by the oxidation of anthropogenic sulfur dioxide (SO_2) emissions, mainly from coal burning. From a climate-change viewpoint, it is SO_4 aerosols that are currently believed to be the most important. They affect the climate in quite complex (and still imperfectly understood) ways.

Under clear sky conditions, sulfate aerosols reflect incoming solar radiation back into space and so perturb the energy balance. More aerosols means more energy reflected back into space and, hence, a cooling. Sulfate aerosols also affect clouds, since the aerosols can act as nuclei for cloud droplets to condense on. More nuclei means more and smaller cloud droplets, which, since clouds with smaller droplets are more reflective ("whiter"), also leads to a cooling. These two aerosol effects are commonly referred to as the direct and indirect aerosol forcing effects. There are other indirect effects, which complicates the issue. For example, changing cloud droplet sizes can change the lifetimes of clouds, which, in turn, can affect the mean amount of cloudiness and the total amount of solar radiation reflected back into space.

On timescales of decades to centuries, solar effects appear to be the most important natural external forcing factor. Explosive volcanic eruptions provide another example of natural external forcing, but these lead mainly to short-term (1-3 year) influences. Anthropogenic forcings occur through increasing greenhouse-gas concentrations and through changes in the atmosphere's aerosol loading, the latter arising primarily as a result of the emissions of sulfur dioxide.

A3 Responses to External Forcing

When external forcing is imposed, the climate system responds by changing the temperature of the atmosphere. As a consequence, all other aspects of the climate system will change — precipitation amounts and patterns, storm tracks and winds, humidity, etc.

The magnitude of these changes is controlled both by the amount of external forcing and, as noted above, by feedback mechanisms (amplifying or moderating effects) within the climate system. The most fundamental of these is water vapor feedback. Any forcing increase will cause warming, which leads to increased evaporation and, hence, to more water vapor in the atmosphere. Since water vapor is an important greenhouse gas, this leads to yet more warming in the lower atmosphere.

The magnitude of this positive feedback depends on how the increased water vapor is distributed within the atmosphere. It is possible to redistribute the water vapor in a way that minimizes the feedback. Lindzen (1990, 1994), for example, has argued that increasing greenhouse-gas concentrations could lead to a drying of the middle to upper troposphere, which would in turn lead to a much smaller water vapor feedback. Inamder and Ramanathan (1998) have tested Lindzen's hypothesis using a wide range of observational data sets. They conclude that the "drying hypothesis of Lindzen (1990) does not explain tropical or global scale changes in water vapor and the atmospheric greenhouse effect in the present atmosphere. By deduction, its validity for the global warming problem is in doubt" (Inamder and Ramanathan, 1998, p. 32193). In other words, observational data simply do not support Lindzen's hypothesis.

There are other important feedbacks involving clouds, snow, and ice extent, etc., most of which have somewhat uncertain magnitude. As a consequence, the magnitude of climate change in response to external forcing, even at the global scale, is quite uncertain. This uncertainty is most simply expressed as an uncertainty in the “climate sensitivity,” an index of how strongly the climate system responds to a change in external forcing. The climate sensitivity is usually expressed in terms of the global-mean, near-surface temperature change that would eventually occur if we doubled the amount of CO_2 in the atmosphere ($\Delta T2x$). The standard uncertainty range for $\Delta T2x$ is 1.5-4.5°C, endorsed by the Intergovernmental Panel on Climate Change (IPCC) in 1990 (Mitchell et al., 1990) and subsequently. In fact, this has been the standard range for many years — what has changed over the years is the confidence interval that it represents. Currently, the range represents roughly the 90 percent confidence interval (see, e.g., Morgan and Keith, 1995) — i.e., expert judgment concludes that there is roughly a 5 percent probability that $\Delta T2x$ is less than 1.5°C, and a 5 percent probability that it is above 4.5°C.

The climate sensitivity concept refers to the eventual (or equilibrium) warming that would occur in response to a specified forcing increase. What would actually happen if forcing were to be applied instantaneously to the system? To understand this, a motoring analogy is useful. First, the sensitivity range, $\Delta T2x$ = 1.5-4.5°C, is akin to a range of vehicle types from, e.g., a Volkswagen (low sensitivity — in other words, a slowish top speed) to a Porsche (high sensitivity, therefore a somewhat higher top speed). Suppose we now start driving the vehicle from a standing start. When we put our foot on the accelerator (i.e., apply external forcing), the car doesn’t immediately leap to top speed. Instead, because we must overcome the mass (inertia) of the car, the car accelerates relatively slowly to top speed. Similarly, when we force the climate system, the response is relatively slow because of the vast thermal inertia of the oceans.

In the car, the speed of the response depends on how rapidly and how far we depress the accelerator, how massive the car is, and how powerful the car is. In the climate system, the response depends on the amount and rate at which the forcing is applied (e.g., the rate of increase of CO_2 concentration), the magnitude of the thermal inertia represented by the oceans, and the climate sensitivity. The analogy also works in reverse. If we take our foot off the accelerator and apply the brake, the car will continue to move forward for some time. Similarly, removing any climate forcing (e.g., by halting the increase in CO_2 concentration) will not immediately halt the increase in global-mean temperature — it may take decades or even centuries before the climate restabilizes. Just as one is committed to moving forward a considerable distance in a moving vehicle, so we are committed to considerable changes in climate even if we could instantaneously stabilize the composition of the atmosphere (which, of course, we cannot!).

A4 Internal Variability of the Climate System

Climate does not change only in response to external forcing. In the absence of such forcing, the heat content of the system would stay constant. However, heat may redistribute itself geographically, between different reservoirs (e.g., between the oceans and the atmosphere), or between different thermodynamic states (such as water vapor to liquid water to ice). Such redistributions occur continuously on timescales from seconds to millennia. In their manifestation as changes in the surface climate of the Earth, they are referred to as internal climatic variability (or, more correctly, unforced internally generated climatic variability).

A5 Summary of Causes of Climate Change

Climate variations may be divided into three types: (1) internal variability, (2) natural externally forced variability, and (3) anthropogenic externally forced variability. These are not mutually exclusive categories — it is possible that part of the climate system's response to external forcing could be through a change in the character of its internal variability.

An important example of internal variability is the El Niño/Southern Oscillation (ENSO) phenomenon. ENSO, which arises from interactions between the ocean and atmosphere in the tropical Pacific, has clear regional consequences over a much wider area, especially for extreme events (see, e.g., Ropelewski and Halpert, 1987). ENSO operates in a quasi-cyclic manner on a time scale of 2-8 years. There is some suggestion that the character of the ENSO phenomenon might change as a result of human activities (sun, 1997; Trenberth and Hoar, 1997). If so, this would be a good example of the potentially blurred distinction between "external forcing" and "internal variability."

The primary natural external forcing factors are changes in solar output and the effects of explosive volcanic eruptions. Volcanic eruptions affect the climate by changing the reflectivity and radiative absorbing properties of the Earth-atmosphere system, through the injection of dust and the production of reflective aerosols in the upper atmosphere. Since the lifetime of these products is only a few years, volcanoes have only short-term effects on the climate. Major eruptions can lead to global-mean cooling of up to 0.5°C, but the cooling lasts for only a few years. For anthropogenic external forcing, the dominant influences are from greenhouse gases and aerosols, although changes in the character of the land surface through land-use changes may be important at the regional (subcontinental) scale.

In climate-change studies, both in studies that attempt to understand the past and those that attempt to predict the future, the critical issue is the relative importance of these three factors. Their relative importance depends on the spatial scale being considered. Over the 20th century, at the global scale, they appear to have been of roughly equal importance. In the future, anthropogenic factors will become increasingly dominant — but the other factors will not disappear, and they will continue to act as important modulators of the human component. At scales smaller than global, the relative importance of the three factors changes, mainly because the magnitude of internally generated variability increases substantially as the spatial scale reduces. For example, the year-to-year variability of temperatures at specific locations is far greater than the corresponding variability of global-mean temperature. As a consequence, while we may even now be able to identify the human component of global-mean temperature change above the "noise" of natural variability, we cannot yet confidently identify the human component on small (sub-continental) spatial scales.